AN IMPORTANT OCEAN FEATURE OVERLOOKED IN CURRENT EL NINO-SOUTHERN OSCILLATION THEORIES

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1. INTRODUCTION

Almost all the theories about the phenomenon of El Nino and Southern Oscillation (ENSO) assume a positive-only correlation in the ocean heat equation between the time rate change of sea surface temperature (SST) anomaly ($\partial \hat{T}_s/\partial t$) and the upper ocean thickness fluctuation ($\hat{h}$). It is very important to justify the validity of this positive-only correlation before developing complicated coupled ocean-atmosphere models.

Based on classical Kraus-Turner type ocean mixed layer (OML) theory, it is found that the positive-only correlation between $\partial \hat{T}_s/\partial t$ and $\hat{h}$ is due to overlook of an ultimately important feature in the OML: **non-upward** turbulent heat flux at the OML base when the OML temperature is higher than the temperature below the OML, as it usually is in the tropics. With consideration of this feature, both positive and negative correlations between the time rate change of SST anomaly and OML depth fluctuation have been found. Furthermore, if initial OML depth is greater than the Monin-Obukhov length-scale (ℓ), the OML quickly shallows to the Monin-Obukhov length-scale, and the negative correlation exists; however, if initial OML depth is smaller than the Monin-Obukhov length-scale, the positive correlation appears. Shift of positive-only correlation to positive/negative correlation will lead to a new ENSO theory.

2. SST EQUATION IN CURRENT ENSO THEORIES

In the current ENSO theories containing ocean thermodynamics, almost all imply a positive-only correlation between time rate change of SST anomaly and OML depth fluctuation. Let us investigate the two typical (Hirst-type and Zebiak-Cane type) heat equations.

2.1 Hirst type heat equation

An oceanic heat equation often appeared in simple coupled ocean-atmosphere models (e.g., Hirst, 1986; Hirst and Lau, 1990) was first presented by Hirst (1986)

$$\frac{\partial \hat{T}_s}{\partial t} = -\overline{T}_x \hat{u} + K_T \hat{h} - \alpha_s \hat{T}_s \qquad (1)$$

where $\hat{T}_s, \hat{h}, \hat{u}$ are perturbations of SST, upper ocean thickness, and zonal currents; $\overline{T}_x$ is the mean zonal temperature gradient; α_s is thermal dissipation coefficient; and $K_T(>0)$ is SST-thickness coupling coefficient. The positive value of K_T, caused by the only appearance of one of the two cases of the OML (i.e., the OML thickness less than the Monin-Obukhov length-scale), guarantees the positive-only correlation between $\partial \hat{T}_s/\partial t$ and $\hat{h}$.

As pointed by Xie et al (1989), the SST equation of Anderson and McCreary (1985)'s coupled air-ocean model also belongs to this category.

2.2 Zebiak-Cane type heat equation

In several coupled numerical models (e.g., Zebiak and Cane 1987; Battisti, 1988; Battisti and Hirst, 1989; Neelin 1991), the following SST anomaly equation is often used:

$$\frac{\partial \hat{T}_s}{\partial t} = -C_T - \Delta(\overline{w} + \hat{w}) \frac{r(\hat{T}_s - T_{\text{sub}})}{H_1}$$
$$- [\Delta(\overline{w} + \hat{w}) - \Delta(\overline{w})] \times \overline{T}_z - \alpha_s \hat{T}_s \qquad (2)$$

where C_T is the horizontal temperature advection; H_1 is the mean upper layer thickness; $\overline{w}, \hat{w}$ are mean and fluctuation of vertical velocity; and T_{sub} is the subsurface temperature, which is determined empirically by

Report Documentation Page

Form Approved
OMB No. 0704-0188

1. REPORT DATE **1992**	2. REPORT TYPE	3. DATES COVERED **00-00-1992 to 00-00-1992**

4. TITLE AND SUBTITLE **An Important Ocean Feature Overlook in Current El Nino-Southern Oscillation Theories**	5a. CONTRACT NUMBER 5b. GRANT NUMBER 5c. PROGRAM ELEMENT NUMBER
6. AUTHOR(S)	5d. PROJECT NUMBER 5e. TASK NUMBER 5f. WORK UNIT NUMBER
7. PERFORMING ORGANIZATION NAME(S) AND ADDRESS(ES) **Naval Postgraduate School,Department of Oceanography,Monterey,CA,93943**	8. PERFORMING ORGANIZATION REPORT NUMBER
9. SPONSORING/MONITORING AGENCY NAME(S) AND ADDRESS(ES)	10. SPONSOR/MONITOR'S ACRONYM(S) 11. SPONSOR/MONITOR'S REPORT NUMBER(S)

12. DISTRIBUTION/AVAILABILITY STATEMENT
Approved for public release; distribution unlimited

13. SUPPLEMENTARY NOTES
Third Conference on Global Change, American Meteorological Society, 52-59

14. ABSTRACT

15. SUBJECT TERMS

16. SECURITY CLASSIFICATION OF:			17. LIMITATION OF ABSTRACT	18. NUMBER OF PAGES	19a. NAME OF RESPONSIBLE PERSON
a. REPORT **unclassified**	b. ABSTRACT **unclassified**	c. THIS PAGE **unclassified**	**Same as Report (SAR)**	**8**	

Standard Form 298 (Rev. 8-98)
Prescribed by ANSI Std Z39-18

$$T_{sub} = T_1[\tanh[b_1(\bar{h}+\hat{h})] - \tanh(b_1\bar{h})], \quad h > 0$$

$$T_{sub} = T_2[\tanh[b_2(\bar{h}-\hat{h})] - \tanh(b_2\bar{h})], \quad h < 0 \tag{3a}$$

where the coefficients in Zebiak-Cane model take the following values: $r = 0.7$, $T_1 = 28^\circ C$, $T_2 = -40^\circ C$, $b_1 = (80m)^{-1}$, and $b_2 = (33m)^{-1}$. Battisti (1988) and Battisti and Hirst (1989) used a little different form for T_{sub}:

$$T_{sub} = a(\bar{h})h, \quad a(\bar{h}) > 0 \tag{3b}$$

Recently, Neelin (1991) adopted a linear form for T_{sub}:

$$T_{sub} = \alpha_0 + \alpha_1 h \tag{3c}$$

where α_0 and α_1 are two empirical constants. The function $\Delta(x)$ in (2) is defined by

$$\Delta(x) = 0, \quad x \le 0$$

$$\Delta(x) = x, \quad x > 0 \tag{4}$$

Utilization of any one of (3a,b,c) with (2) leads to a positive-only correlation between $\partial\hat{T}_s/\partial t$ and $\hat{h}$, i.e., the increase (or decrease) of $\hat{h}$ causes the increase (or decrease) of $\partial\hat{T}_s/\partial t$. **Is this positive-only correlation correct?** The answer is **no.** Recently, Chu (1991a,b, c) pointed out the possibility of existence of **negative correlation** between $\partial\hat{T}_s/\partial t$ and $\hat{h}$, base on the surface wind condition. In this note, a further investigation of thermodynamical and dynamical structure of the OML explores the oceanic condition for the **negative correlation.**

3. OCEAN MIXED LAYER MODEL

Both atmospheric and oceanic planetary boundary layers are ultimately important for ocean-atmosphere interaction since fluxes of momentum and buoyancy across the air-ocean interface are major driving forces (or dampening factors) for the two fluids. Therefore, before designing any ENSO model (either theoretical or numerical), whether the boundary layer physics is reasonable or not should be first checked up.

Kraus and Turner (1967) developed a OML model by considering the layer to be vertically homogeneous: heat inputs to the layer and mass entrained at the bottom of the layer are instantaneously mixed uniformly through the layer. Turbulence in the upper ocean is strong to maintain a vertically uniform temperature down to a certain depth, h. The OML temperature (T_s) and depth (h) are important variables predicted by OML models. Below the depth h, there is a rapid decrease in temperature across what is called the entrainment zone. The thickness of entrainment zone ε is assumed infinitesimally small. Turbulence is assumed to be zero below the entrainment zone, where the vertical velocity w is determined by the dynamics of the ocean interior. The model OML is depicted in Fig.1. Vertically uniform temperature in the OML (therefore, no vertical advection in the OML) and no turbulence in the deeper layer below the entrainment zone lead to three different forms of heat equation for three parts of the ocean:

$$\frac{\partial T_s}{\partial t} + \frac{\partial}{\partial z}(\overline{w'T'}) = \frac{\partial F}{\partial z}, \quad z \ge -h \tag{5a}$$

$$\frac{\partial T}{\partial t} + w\frac{\partial T}{\partial z} + \frac{\partial}{\partial z}(\overline{w'T'}) = \frac{\partial F}{\partial z}, \quad -h > z \ge -h-\varepsilon \tag{5b}$$

$$\frac{\partial T}{\partial t} + w\frac{\partial T}{\partial z} = \frac{\partial F}{\partial z}, \quad z < -h-\varepsilon \tag{5c}$$

where $\overline{w'T'}$ is vertical turbulent heat flux; ρ_{w0} is the characteristic value of sea water density; c_{pw} is specific heat of sea water under constant pressure, and F is downward flux of solar radiation in the ocean computed by

$$F = F_0 e^{\gamma z}, \quad F_0 = \frac{I_0}{\rho_{w0} c_{pw}} \tag{6}$$

where I_0, γ are the solar radiation incident on the ocean surface and the average extinction coefficient. Vertical integration of heat equation (5) from the ocean surface to the OML base leads to

$$h\frac{\partial T_s}{\partial t} = (\overline{w'T'})_{-h} - (\overline{w'T'})_0 + F_0 - F_{-h} \tag{7}$$

At the ocean surface, the turbulent heat flux must equal the net heat transfer through the air-ocean interface, i.e.,

$$(\overline{w'T'})_0 = \frac{R_b + H_s + LE}{\rho_{w0} c_{pw}} \tag{8}$$

where R_b is the net heat loss by longwave radiation from the sea surface; and H_s and LE are the upward sensible and latent heat fluxes across the air-ocean

interface. Vertically integrated turbulent kinetic energy (TKE) equation is given by (Kraus and Turner, 1967)

$$\int_{-h}^{0} \overline{w'T'}\,dz = -(G-D) \tag{9a}$$

where D denotes the dissipation within the OML and G the TKE input from the wind. The simplest parameterization for $(G-D)$ is (Denman, 1973)

$$G-D=\frac{m(C_D\rho_{a0}/\rho_{w0})^{3/2}|\vec{V}_a|^3}{g\alpha}>0 \tag{9b}$$

where $\vec{V}_a$ is the horizontal wind at 10 m height; ρ_{a0} is the characteristic value of surface air density; C_D is the drag coefficient; g is gravity; α is thermal expansion coefficient of sea water; and m is proportionality. Integration of the heat equation (5) with respect to z from a depth z inside the OML ($z>-h$) to the ocean surface leads to

$$-z\frac{\partial T_s}{\partial t}+(\overline{w'T'})_0-(\overline{w'T'})_z=F_0-F(z) \tag{10}$$

Further integration of (10) with respect to z from the OML base to the ocean surface gives rise to

$$\frac{h^2}{2}\frac{\partial T_s}{\partial t}=\int_{-h}^{0}\overline{w'T'}\,dz-h(\overline{w'T'})_0+h(F_0+\frac{1}{h}\int_{-h}^{0}F\,dz) \tag{11}$$

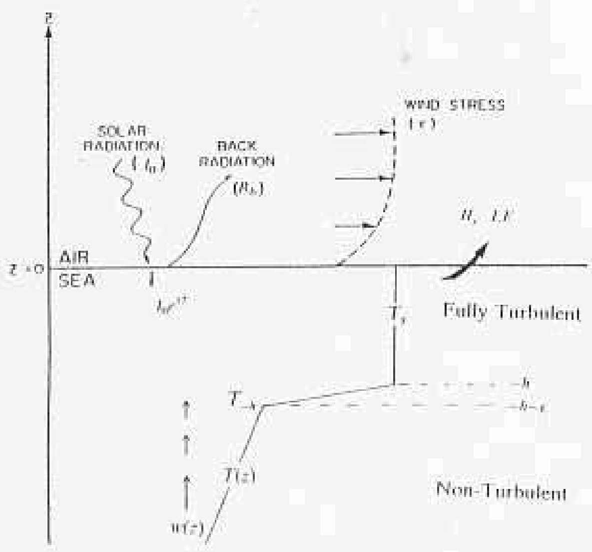

Fig.1. Ocean mixed layer model.

Elimination of $\partial T_s/\partial t$ from (5) and (11) and utilization of (9a) leads to a relation for computation of turbulent heat flux at the OML base

$$(\overline{w'T'})_{-h}=-(\overline{w'T'})_0-\frac{2(G-D)}{h}+2[\frac{F_0}{2}(1+e^{-\gamma h})-\frac{F_0}{\gamma h}(1-e^{-\gamma h})] \tag{12}$$

If the penetration depth of solar radiation is much smaller than the OML thickness, i.e., $\gamma h \gg 1$, the exponent $e^{-\gamma h}$ is negligible against 1, and therefore Eq.(12) can be simplified as

$$(\overline{w'T'})_{-h}=-2[\frac{1}{2}(\overline{w'T'})_0+\frac{(G-D)}{h}-F_0(\frac{1}{2}-\frac{1}{\gamma h})] \tag{13}$$

The turbulent heat flux at the OML base is either downward or zero, never upward, i.e.,

$$(\overline{w'T'})_{-h}\le 0 \tag{14}$$

due to higher OML temperature (T_s) versus deeper layer temperature (Fig.1). With consideration of this constrain, a better form is recommended to replace (13)

$$(\overline{w'T'})_{-h}=-\frac{Q_0}{\rho_{w0}c_{pw}}\Delta(\frac{\ell-h}{h}) \tag{15}$$

where Q_0 is the net downward heat flux plus radiation at the ocean surface:

$$Q_0=\rho_{w0}c_{pw}[F_0-(\overline{w'T'})_0] \tag{16}$$

which is generally positive in the tropics when the low frequency mode is concerned; and ℓ is Monin-Obukhov length scale defined by

$$\ell=\frac{2(G-D+F_0/\gamma)}{Q_0/\rho_{w0}c_{pw}} \tag{17}$$

Integrating the heat equation (5b) across the entrainment zone from $z=-h-\varepsilon$ to $z=-h$ and taking the limit as $\varepsilon\to 0$, a jump condition has been obtained (Denman, 1973)

$$(\overline{w'T'})_{-h}=-\Delta(w_{-h}+\frac{\partial h}{\partial t})(T_s-T_{-h}) \tag{18}$$

where w_{-h}, T_{-h} are vertical velocity and temperature at the base of the entrainment zone, respectively.

In fact,

$$w_e = w_{-h} + \frac{\partial h}{\partial t}$$

is called the entrainment velocity which were used in some ENSO models containing the OML (e.g. Anderson and McCreary, 1985; Hirst, 1986; Xie et al, 1989). Since the OML temperature is higher than the temperature below the OML, the function $\Delta(x)$ also guarantees **non-upward** turbulence heat flux at the OML base. If $w_{-h} + \partial h/\partial t > 0$, there is entrainment mixing (causing downward turbulent heat flux) at the base of the entrainment zone. If $w_{-h} + \partial h/\partial t \leq 0$, there is no entrainment mixing (causing zero turbulent heat flux) at the base of the entrainment zone. Eqs.(7), (15), (17), and (18) are basic equations for Kraus-Turner type OML models. Eq.(15) is used to compute turbulent heat flux $\overline{(w'T')}_{-h}$, Eq.(7) is used to determine the OML temperature T_s, and Eq.(18) is used to calculate time rate change of OML depth dh/dt in entrainment mixing case, e.g.,

$$\frac{\partial h}{\partial t} = -w_{-h} - \frac{\overline{(w'T')}_{-h}}{(T_s - T_{-h})}, \qquad (h < \ell) \quad (19)$$

From Wyrtki's (1981) estimation, the mean upwelling velocity at the base of the Ekman layer ($z = -\delta_E$) is nearly $1\ m/day$. Theoretically, the Ekman depth at the equator is infinity. If $\delta_E \simeq 200\ m$ is thought to be a reasonable estimation of Ekman layer thickness, the mean vertical velocity at the mean mixed layer depth can be roughly evaluated by

$$w_{-\ell} \simeq \frac{\ell}{\delta_E} w_{-h} \simeq 0.1\ m/day$$

where the mean mixed layer depth, taken as 24 m, is from the data collected during Hawaii-to-Tahiti Shuttle Experiment (Schneider and Muller, 1990).

Fig.2 indicates the OML deepening rate caused by the turbulent heat flux at the OML base for five different values of the temperature jump at the OML base $(T_s - T_{-h})$. For h not very close to ℓ, in the righthand side of (19) the second term is much larger than the first term, which means that the change of OML depth is more likely by the strength of turbulent heat flux at the OML base rather than by the Ekman pumping. Eq.(9b) indicates that an increase of surface wind speed leads to an increase of $G - D$, which in turn augments the OML deepening rate through the increase of the downward turbulent heat flux at the OML base, $-\overline{(w'T')}_{-h}$, as $h < \ell$ [*see*(13)] ; or increases the OML depth as $h = \ell$ [*see*(17)]. In other words, the increase of surface wind speed generally increases the OML depth.

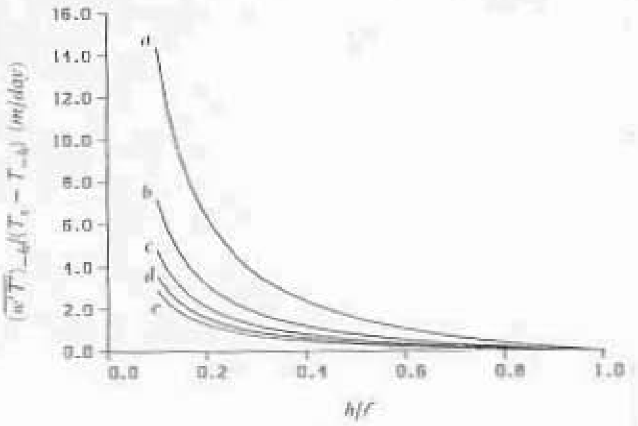

Fig.2. Dependence of OML deepening rate (for $h < \ell$) caused by the turbulent heat flux at the OML base, $\overline{(w'T')}_{-h}/(T_s - T_{-h})$, on h/ℓ for different values of $(T_s - T_{-h})$: (a) 1°C, (b) 2°C, (c) 3°C, (d) 4°C, and (e) 5°C.

4. AN IMPORTANT FEATURE OVERLOOKED IN THE ENSO THEORIES

Another form of OML heat equation can be obtained by the substitution of (15) into (7):

$$\left[\frac{Q_0}{\rho_{w0} c_{pw} \mathscr{H}}\right]^{-1} \frac{\partial T_s}{\partial t} = \frac{\mathscr{H}}{h}\left[1 - \Delta\left(\frac{\ell - h}{h}\right)\right] \quad (20)$$

Here, $\mathscr{H}$ is the characteristic OML thickness. The time rate change of SST is determined by the Monin-Obukhov length-scale ℓ and OML depth h

An ultimately important feature has been overlooked in the current ENSO theories is the existence of two different analytical forms of the OML heat equation depending on the initial OML depth. If the initial OML depth is greater than the Monin-Obukhov length-scale ($h > \ell$), the computed value of the turbulent heat flux at the OML base $\overline{(w'T')}_{-h}$, which is only determined by the ocean surface wind forcing $(G - D)$, OML depth (h), and the surface net heat flux [*see*(13)], will be positive, which is against the physical law that there is no **upward** turbulent heat flux at the OML base when $T_s > T_{-h}$. Moreover, the function $\Delta[(\ell - h)/h] = 0$, which makes

$$(\overline{w'T'})_{-h} = 0$$

and turns the SST equation (20) into

$$\left[\frac{Q_0}{\rho_{w0} c_{pw} \mathscr{H}}\right]^{-1} \frac{\partial T_s}{\partial t} = \frac{\mathscr{H}}{h}, \qquad (h = \ell) \quad (21a)$$

which shows that the time rate change of SST ($\partial T_s/\partial t$) monotonically decreases with the increase of h (Fig.3). It is easy to find in this case that the increase of the OML depth h causes the decrease of the time rate change of SST, which implies a **negative correlation** between the time rate change of SST anomaly and OML depth fluctuation, which was overlooked in the current ENSO models.

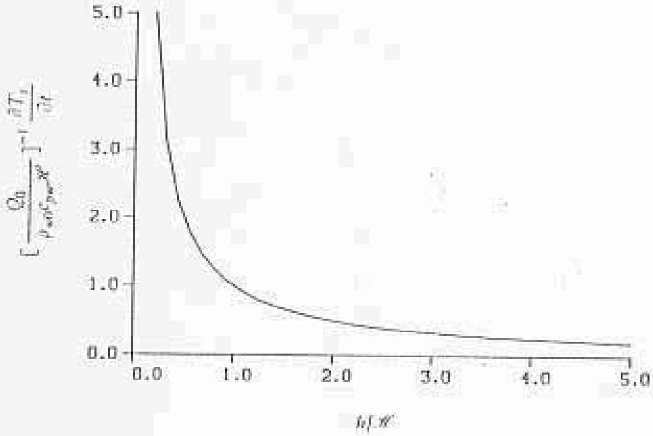

Fig.3. Time rate change of SST versus OML thickness for $h < \ell$.

As the OML depth is smaller than the Monin-Obukhov length scale ($h < \ell$), i.e., the turbulent heat flux at the OML base is negative

$$(\overline{w'T'})_{-h} < 0$$

and the OML heat equation (20) becomes

$$\left[\frac{Q_0}{\rho_{w0} c_{pw} \ell}\right]^{-1} \frac{\partial T_s}{\partial t} = \frac{\ell}{h}\left(2 - \frac{\ell}{h}\right), \quad (h < \ell)(21b)$$

which shows that the time rate change of SST ($\partial T_s/\partial t$) monotonically increases with the increase of h (Fig.4). This leads to a **positive correlation** between the time rate change of SST anomaly and the OML depth fluctuation, which has been well represented in the current ENSO theories.

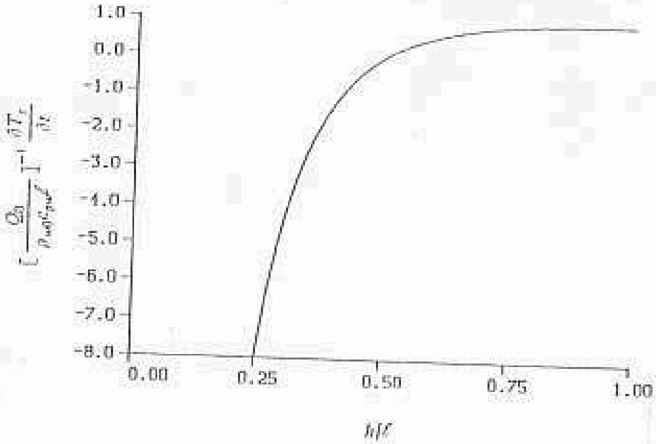

Fig.4. Time rate change of SST versus OML thickness for $h > \ell$.

5. GENERATION OF TWO TYPES OF UNSTABLE MODES

Since the OML heat equation has two different forms (21a) and (21b) due to initial OML depth versus the Monin-Obukhov length scale, it is anticipated that the features of unstable modes generated by coupled air-ocean system should be different.

The tropical subcloud layer is usually dominated by mean easterlies. The low-level atmospheric convergence and convection ("Convection A") are generally associated with the warm water (Gill, 1980). In the east (or west) of the warm water region, the surface wind speed is enhanced (or reduced), and therefore, the OML depth h is increased (or decreased). There are two main cases which are illustrated as follows.

5.1 Small initial OML thickness ($h < \ell$)

In this case, the SST equation is given by (21b), showing a positive correlation between $\partial T_s/\partial t$ and the OML depth (Fig.3). Therefore, a positive OML fluctuation ($\hat{h} > 0$) leads to a positive time rate change of SST anomaly ($\partial \hat{T}_s/\partial t > 0$). In the east (or west) of Convection A, the OML depth fluctuation is positive (or negative), which in turn gives rise to a positive (or negative) time rate change of SST anomaly (Fig.5). Thus, positive SST anomaly and associated unstable convective disturbances (denoted by "Convection B") will appear in the east of the initial warm water region. Applying the same argument to Convection B, positive

SST anomaly and associated unstable convective disturbances (denoted by "Convection C") will appear in the east of Convection B (Fig.6) due to the further increase of OML depth caused by the enhanced surface easterlies (sum of the background easterlies, easterlies associated with Convections A and B). Unstable convective disturbances are unlikely to appear between Convection A and Convection B due to the opposite directions of the surface winds associated with A and B. Therefore, as $h < \ell$, the unstable mode is generated in the east of the initial warm water region and propagates eastward.

For a typical La Nina condition, atmospheric deep convection (Convection A) develops over the western pacific warm water. A shallow and cool OML water occupies the eastern Pacific. The OML thickness is generally satisfied the condition $h < \ell$. The warm SST disturbances and associated atmospheric convection are generated in the east of Convection A and propagate eastward, which leads to a eastward propagation of the surface westerlies (westerlies associated with Convections A, B, and C). At the same time, the OML thickness h increases. The whole process features the transition from La Nina to El Nino.

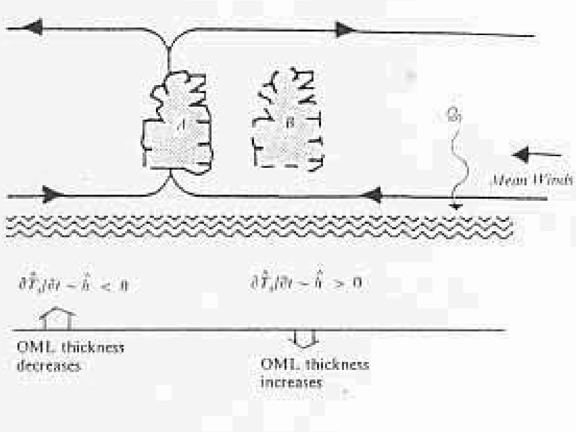

Fig.5. Positive SST anomaly and associated convective disturbances (Convection B) generated in the east of Convection A ($h < \ell$).

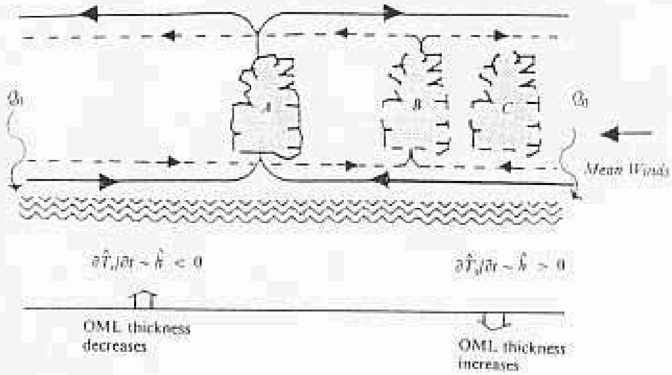

Fig.6. Eastward propagation of positive SST anomaly and associated convective disturbances ($h < \ell$).

5.2 <u>Large initial OML thickness ($h \geq \ell$)</u>

In this case, the OML quickly shallows to the Monin-Obukhov length-scale. The OML shallowing time scale is estimated by (Garwood, 1977)

$$\tau = \frac{h}{<\overline{E}>^{1/2}}$$

where $\overline{E}$ is the mean TKE in the OML. If $<\overline{E}>^{1/2} \sim 10\, cm/s$ and $h \sim 100\, m$, the OML shallowing time scale (τ) is nearly 1000 seconds, which is so short comparing to the ENSO time scale. Thus, it is reasonable to assume that the OML instantly shallows to the Monin-Obukhov length scale ℓ when the initial OML depth is greater than ℓ. The SST equation is given by (21a), showing a negative correlation between $\partial T_s/\partial t$ and the OML depth (Fig.2), which means that a negative OML fluctuation ($\hat{h} < 0$) leads to a positive time rate change of SST anomaly ($\partial \hat{T}_s/\partial t > 0$). In the west (or east) of initial Convection D, the OML depth fluctuation is negative (or positive), which in turn gives rise to a positive (or negative) time rate change of SST anomaly (Fig.7). Thus, positive SST anomaly and associated unstable convective disturbances (denoted by "Convection E") will appear in the west of Convection D. Applying the same argument to Convection E, positive SST anomaly and associated unstable convective disturbances (denoted by "Convection F") will appear in the west of Convection E (Fig.8) due to the further decrease of OML depth caused by the reduced surface easterlies (the background easterlies counterbalanced by westerlies associated with Convections D and E). Unstable convective disturbances are un-

likely to appear between Convection D and Convection E due to the opposite directions of the surface winds associated with D and E. Therefore, as $h > \ell$, the unstable mode is generated in the west of the initial warm water region and propagates westward.

For a typical El Nino condition, a deep and warm OML water occupies the eastern Pacific, where the atmospheric deep convection (Convection D) develops. The large OML thickness is quickly adjusted to the Monin-Obukhov length-scale, $h = \ell$. The warm SST disturbances and associated atmospheric convection are generated in the west of Convection D and propagate westward, which leads to a westward propagation of the surface westerlies (westerlies associated with Convections D, E, and F). At the same time, the OML thickness h decreases. The whole process features the transition from El Nino to La Nina.

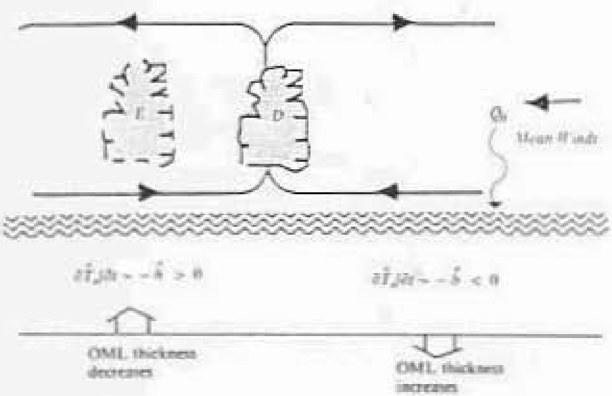

Fig.7. Positive SST anomaly and associated convective disturbances (Convection E) generated in the west of Convection D ($h = \ell$).

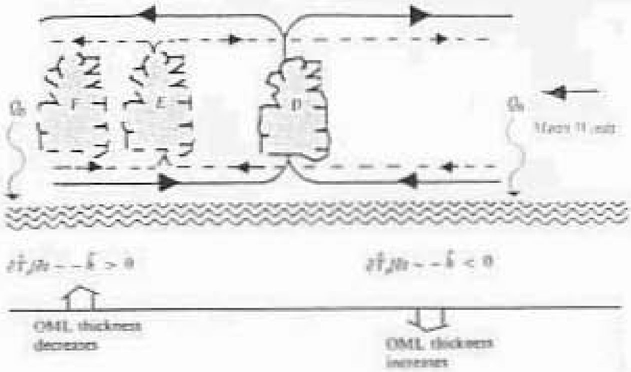

Fig.8. Westward propagation of positive SST anomaly and associated convective disturbances ($h = \ell$).

6. SUMMARY

(1) Both eastward and westward propagating unstable modes described here is produced purely by the SST-OML thickness feedback. Positive (or negative) correlation between $\partial \hat{T}_1/\partial t$ and $\hat{h}$ will generate eastward (or westward) propagating unstable mode.

(2) Shift from positive-only correlation to positive/negative correlation between $\partial \hat{T}_1/\partial t$ and $\hat{h}$ leads to a discovery of a new mechanism of La Nina and El Nino transition based on the magnitude of OML thickness (h) relative to the Monin-Obukhov length-scale (ℓ). It also indicates the existence of an OML switcher for this transition.

(3) The vertical advection term can not explicitly appear in the OML heat equation because it only affects the turbulent heat flux at the OML base through the jump condition (18). As $h < \ell$, the vertical advection changes the OML temperature through the change of OML thickness. As $h \geq \ell$, the OML quickly shallows to the Monin-Obukhov length-scale and the turbulent heat flux at the OML base becomes zero. Therefore, in this case the vertical advection no longer affects the OML.

ACKNOWLEDGEMENTS

This work was funded by the Office of Naval Research and the Naval Postgraduate School. The author thanks R.L. Haney and R.W. Garwood, Jr. at the Naval Postgraduate School for invaluable discussion.

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